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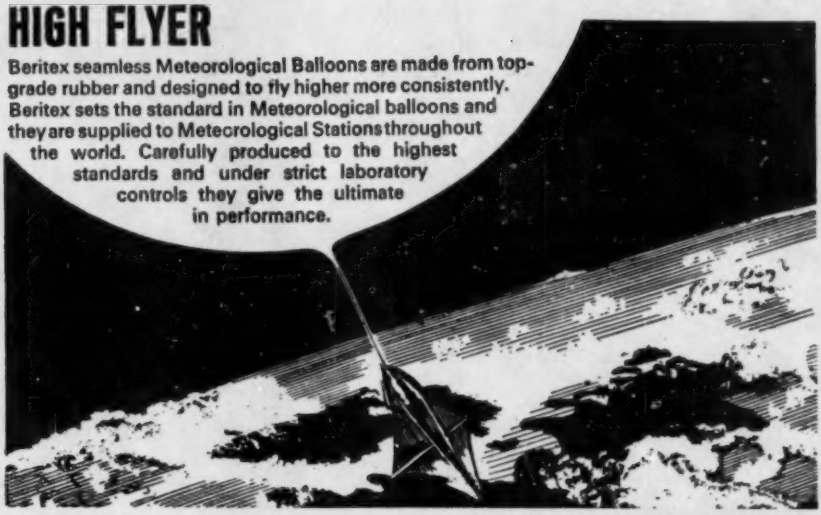

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By S. G. CORNFORD

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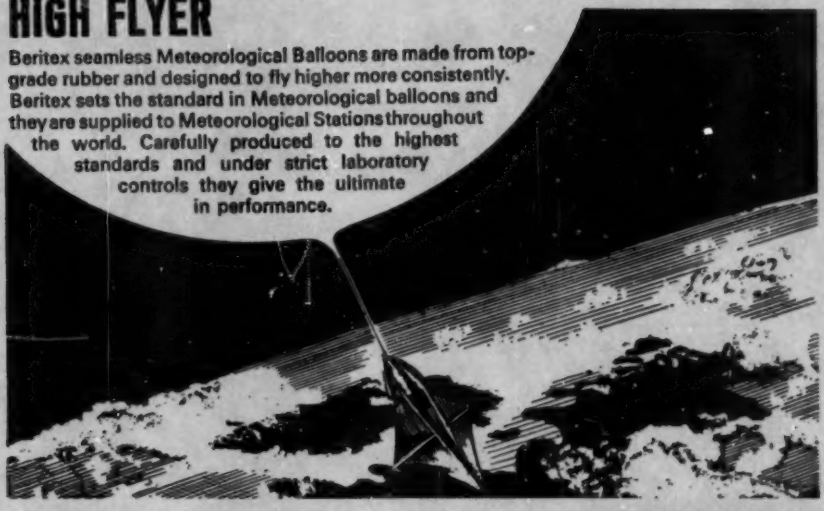

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Instrumentation. Three observers took simultaneous readings of temperature, dew-point, height and airspeed at height intervals of about 30 m. They used a Meteorological Office balanced-bridge flat-plate resistance thermometer, a Dobson-Brewer hygrometer and calibrated aneroid instruments for pressure-height and airspeed.

Changes in air position from that at the beginning of the first helix were monitored using a Royal Air Force air position indicator Mk 1B (API). There is, however, a fundamental difficulty in the concept of air position in sheared or convergent flows. The API continuously monitors the speed and direction of the aircraft's movement through the air (true airspeed and heading), integrates them with respect to time and presents the result as a horizontal displacement from an initial position. Instrument errors and aircraft sideslip apart, this will be a true displacement if the wind has the same velocity all along the aircraft's path. (The path must also be horizontal but normal rates of ascent and descent of the aircraft (say 5 m/s) are very much less than the forward speed (say 75 m/s) so that the horizontal component (74·83 m/s) is very closely equal to the airspeed measured along the sloping flight path. In the present results this source of error may be neglected.) On the present flight, the aircraft was nominally brought back to an initial air position P_o and the same period was spent in each part of the flow. With such a flight pattern the difference, δ, between P_o and the final air position, is the integrated effect of the winds encountered along the flight path and *measured relative to* P_o. Relative winds may arise because of convergence or shear encountered along the flight path. Horizontal convergences with a typical value of 10^{-6}/s will lead to values of δ which are comparable with the accepted error of the API in normal flying use ($\pm$ 2 per cent of the air distance flown). During the present flight there were no indications of abnormally large values of horizontal convergence. On the other hand, shear along the flight path is in general likely to be a more frequent cause of considerable relative winds, especially when the aircraft is ascending or descending as in the present case. On this flight the shear could not be measured but an estimate based on the results of Findlater *et alii*[1] suggests that the difference between corresponding air positions at any two levels probably did not exceed 5 km for each hour flown. As it was impractical to consider the air position at each height separately in order to allow the air position to vary with height, this shear effect was regarded as an error in the mean air position and combined with the other errors already discussed. The initial air position was regarded as a vertical line through P_o. With time this became a circular cylinder of uncertainty with a radius which expanded at about 8 km/h. For practical purposes mean air positions were taken as the position of the vertical axis of the cylinder.

Results. The aircraft's mean track over the ground is shown in Figure 1, together with the synoptic chart for 1800 GMT and the pattern of sea surface temperatures for the period 10–14 February prepared in the Central Forecasting Office, Bracknell. Throughout the flight the sky was clear of cloud. In the south-easterly flow between an anticyclone centred near Riga and a deep depression 500 km west of Shannon, a large area of almost clear skies extended from the North Sea through Belgium and Germany to Italy and the Balkans. A warm front extending from the depression centre to the Isles of Scilly and Brest was associated with rain in the south-west of the United Kingdom. At the time of the flight cirrus extended as far north-east as the coast of Yorkshire but not over the sea in the operating area. Rain reached Yorkshire early in the morning after the flight. A sounding was made near 54°N 1°E (marked S on Figure 1) between 1536 and 1549 GMT. The results of this are shown in Figure 2.

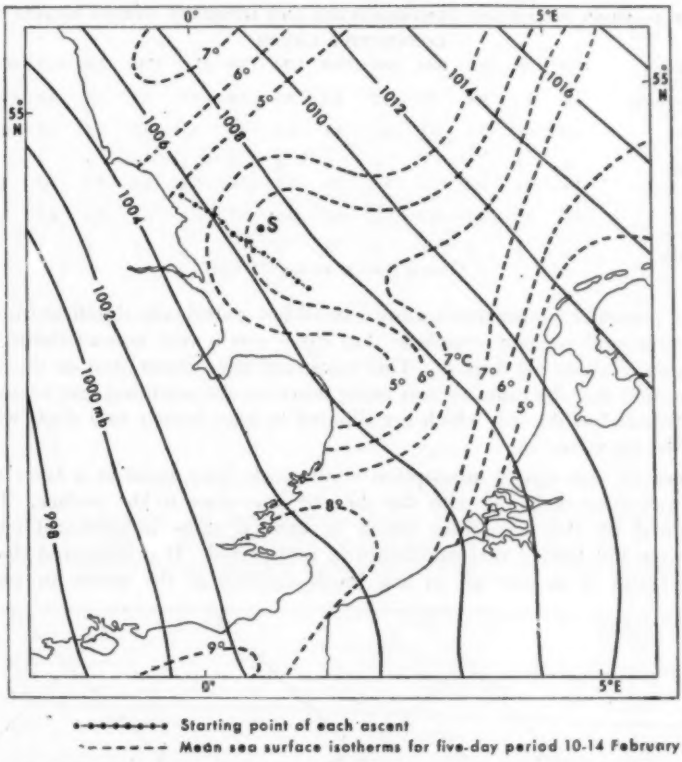

FIGURE 1—SYNOPTIC CHART FOR 1800 GMT, 15 FEBRUARY 1967

S = Sounding made near 54°N 1°E between 1536 and 1549 GMT.
Mean sea surface isotherms are shown for the five-day period 10–14 February.

For each pressure-height, air temperature and dew-point were plotted against distance moved over the sea from the starting point. The resulting pattern of isotherms is shown as Figure 3. Dew-points were expressed as humidity mixing ratios and the pattern of humidity is shown as Figure 4. Air temperatures were expressed as potential temperatures θ. The pattern of isentropes is shown as Figure 5.

The mean winds in the layer examined by the aircraft were measured as 130° 66 km/h for the period 1620 to 1723 GMT and 130° 72 km/h for the period 1723 to 1831 GMT. The vector error in these winds is probably ± 5 km/h.

Discussion.

(i) *Air mass modification.* From the measurements made beneath the inversion, means of potential temperature and humidity mixing ratio have been calculated. For each ascent the mean — and standard deviation — were found of the six or so values between the level below which there were no readings on one or more ascents and the base of the inversion. The results are shown in Table I.

TABLE I—MEAN POTENTIAL TEMPERATURES AND HUMIDITY MIXING RATIOS IN THE CONVECTIVE LAYER

Time of start of ascent GMT	1620	1631	1641	1651	1701*	1710	1718	1728	1737	1746	1756	1805	1816	1825
Mean potential temp. (°C)	4.3	4.5	4.4	4.5	4.5	4.5	4.5	4.6	4.6	4.5	4.7	4.6	5.1	4.8
Standard deviation	0.3	0.2	0.2	0.1	0.2	0.1	0.1	0.1	0.1	0.2	0.1	0.1	0.3	0.1
No. of obs.	6	8	7	7	6	5	6	5	5	5	4	5	5	6
Mean h.m.r. (g/kg)	3.1	3.1	3.2	3.3	3.4	3.3	3.7	3.6	3.2	3.4	3.5	3.3	3.3	3.5
Standard deviation	0.3	0.2	0.2	0.3	0.2	0.2	0.2	0.2	0.1	0.2	0.1	0.1	0.2	0.2
No. of obs.	6	7	6	7	5	4	3	5	4	5	4	5	6	5

*Time of sunset at MSL was 1704 GMT.

The potential temperatures show a small but statistically significant increase with time and we may conclude that there was a real non-adiabatic mean warming of about 0·2 degC/h. This continued after sunset, and on the simple assumption that the same air was being flown in, the warming may reasonably be attributed to the sea, which is estimated to have been 1 to 2 degC warmer than the air at 100 m.

However, this simple assumption is obviously least valid in a layer which is in turbulent exchange with the sheared flow close to the surface. This is illustrated by the humidities which in general show insignificant changes with time but have a real maximum in mid-period. It is suggested that this is the result of moister air at low levels arriving at the mean air position

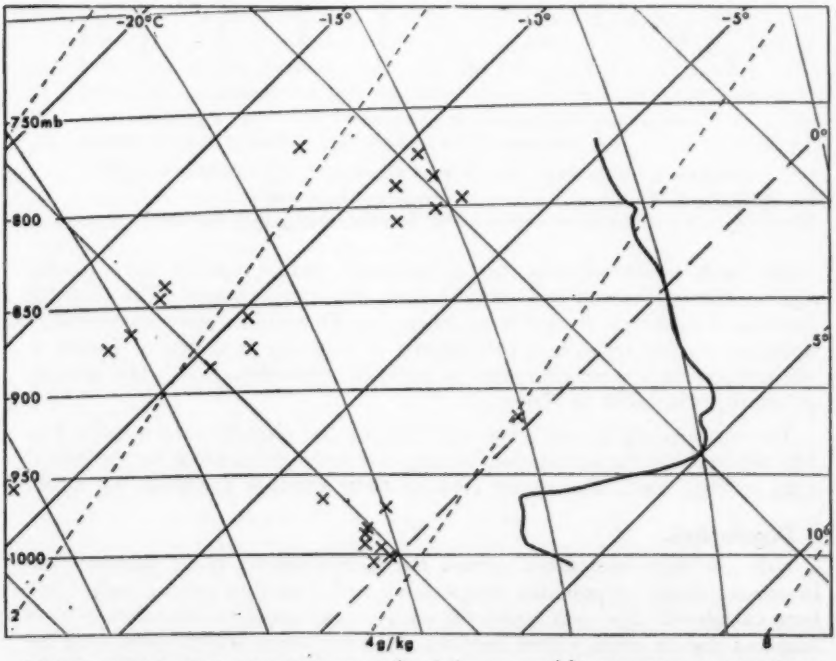

FIGURE 2—SOUNDING MADE NEAR 54°N 01°E, POINT 's' ON FIGURE 1, BETWEEN 1536 AND 1549 GMT, 15 FEBRUARY 1967

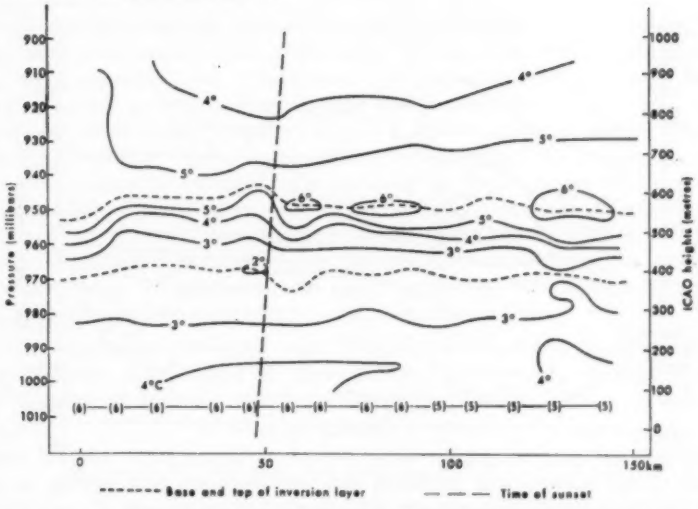

FIGURE 3—VARIATION OF AIR TEMPERATURE WITH PRESSURE-HEIGHT AND HORIZONTAL DISTANCE TRAVELLED BY THE AIR

The figures in brackets are estimated sea surface temperatures plotted at the estimated sea-level pressure.

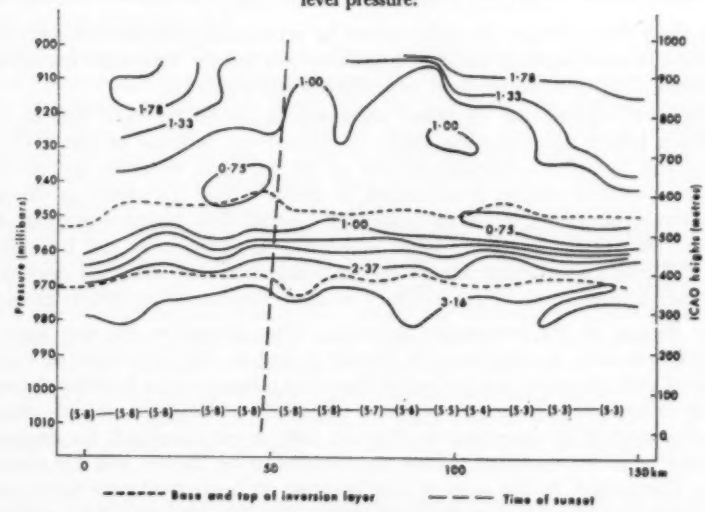

FIGURE 4—VARIATION OF HUMIDITY MIXING RATIO, x, WITH PRESSURE-HEIGHT AND HORIZONTAL DISTANCE TRAVELLED BY THE AIR

Isopleths are drawn at equal intervals of $\log_{10}x$, but are marked with the value of x in grammes per kilogramme.

The figures in brackets are saturation humidity mixing ratios over sea water of 3·5 per cent salinity, plotted at the estimated sea-level pressure.

because of the low-level shear, taking part in the convection and then, in its turn, being replaced by rather drier air again. In a turbulent sheared flow, the simple concept of a trajectory is not valid and the humidity measure-

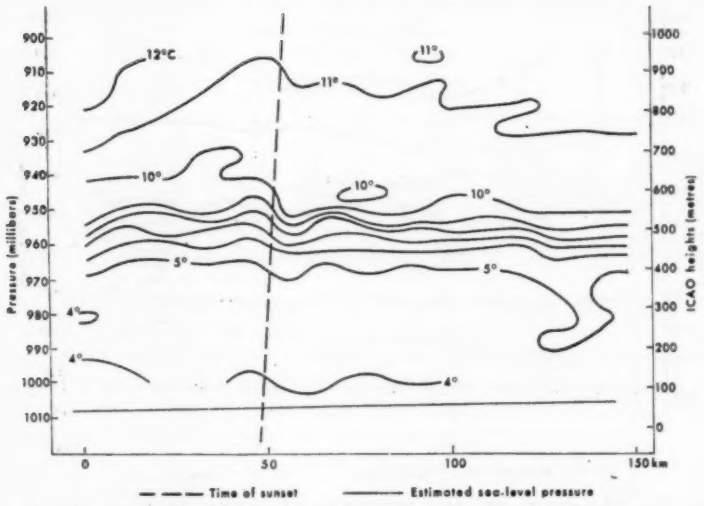

FIGURE 5—VARIATION OF POTENTIAL TEMPERATURE, θ, WITH PRESSURE-HEIGHT AND HORIZONTAL DISTANCE TRAVELLED BY THE AIR

ments show that changes brought about by continuing differential advection with simultaneous vertical mixing can swamp any longer time-scale moistening by the sea, a process which must nevertheless still continue.

A simpler explanation of these observations could be that the aircraft had flown into a neighbouring patch of moister air because of errors in the air position but this is considered not to be so. The error in air position between successive ascents is estimated at about 1 km. In climbing through the 300 m of air beneath the inversion the aircraft flew round approximately one turn of a helix of about 3-km diameter. Consequently the humidities given in Table I are already means on the 3-km scale and unknown deviations of another kilometre or so are unlikely to have produced significant changes.

(ii) *Waves at the temperature inversion.* The sinuosities in the lines of constant humidity mixing ratio in Figure 4 and in the isentropes of Figure 5, suggest that waves with a period of about 20 minutes were running through the air across the inversion surface. Spatial undulations found during horizontal runs in an inversion in clear air have been described, for example, by Grant[2] and in inversions above stratocumulus by Zajcev and Ledohovič[3] and by Cornford.[4] In the present case because no horizontal runs were made there must be extra caution in interpreting the observations. For example, their representativeness on any particular horizontal scale is determined by variations on other scales. It is known that significant differences occur on scales smaller than the 3-km diameter helix flown by the aircraft; Grant's observations, for instance, show temperature changes of up to 4 degC over 1 km, although his was a somewhat extreme case. In the present case, it is the fairly general coherence of the waves in the vertical (especially in Figure 4) which suggests that the large waves with a period of about 20 minutes produced changes greater than any small-scale variations. In ascending

through the inversion the aircraft also moved forward along its helical path some 5 km to the opposite side of the helix. If there were horizontal differences they were included and the general vertical coherence remained despite them. Another possible source of error is the measurement of height, which is common to both the temperature and humidity patterns and so decreases their independence. However, the differences in height of the same isentropic or constant humidity surface at the inversion level are greater than any likely errors in the height measurement, since although absolute pressure-heights measured by aneroid altimeters may suffer from errors of tens of metres, or several millibars, because of lag, yawing of the aircraft and so on, in the present series of measurements the flight conditions were similar throughout so that relative errors should be smaller, 1 to 2 mb or less.

In a comment, published with Grant's observations, Sawyer[2] gives a formula from which may be derived the frequency, n, of short internal gravity waves at a simple density discontinuity. This is

$$n^2 = (g/4\pi\lambda)(\Delta\theta/\theta)$$

where λ is the wavelength, θ the potential temperature, g the acceleration due to gravity and $\Delta\theta$ the change of potential temperature across the discontinuity. In the present example, the frequency is about three per hour, $\theta \simeq 280°K$ and $\Delta\theta \simeq 5°K$. These give a wavelength of about 20 km and a speed of propagation, $n\lambda$, of 60 km/h. As in Grant's case the calculated speed of propagation is about 85 per cent of the wind speed. The direction of motion of the waves is not known.

(iii) *Changes above the inversion.* The significant feature above the inversion is the flat temperature minimum in mid-period above 750 m — see Figure 3. The arguments against this being due to horizontal variability within the air mass are as before. The feature might be the result of differential advection in wind shear but it calls to mind the temperature minima found at sunset during fine weather over land during the Great Plains Turbulence Field Programme (Lettau and Davidson[5]). Figure 6 shows the variation of temperature with time for the four occasions during the Programme on which two-hourly soundings were made for at least six hours before and six hours after sunset. In each case the temperature fell at all levels up to 2 km after an afternoon maximum. Above the evening surface inversion the temperature then recovered, at least temporarily. An examination of the other, shorter, periods of observations around sunset given by Lettau and Davidson shows that the effect is not always as clear cut as Figure 6 may suggest. Nevertheless, the evidence is sufficient to suggest a fairly high frequency of temperature minima at 1 to 2 km over land on clear days around sunset.

The importance of the present observations is that they show a temperature minimum at the same time over the sea, about 300 km downwind of land and above a persistent inversion. They also show that on 15 February 1967 over the North Sea at least, the sunset temperature minimum was the result of a temporary lifting and then sinking of the air above the level of the inversion.

We can come to this conclusion because of the fortunately sharp gradients of humidity in the vertical. Figure 4 shows that the humidity mixing ratios decreased with height to a minimum at the top of the inversion layer and

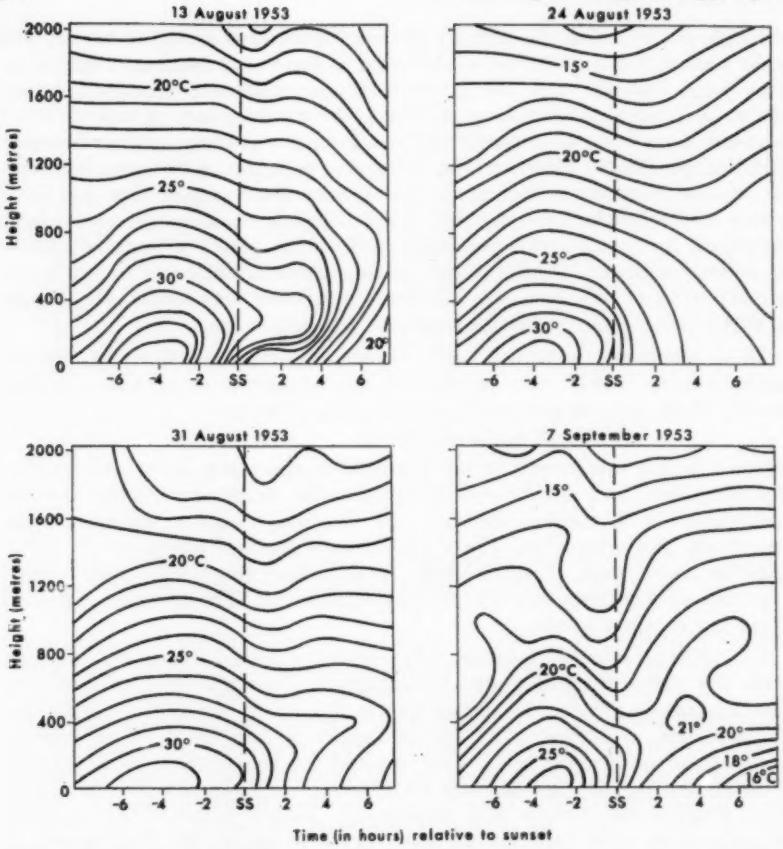

FIGURE 6—TEMPERATURE VARIATION WITH HEIGHT AND TIME FOR AT LEAST SIX HOURS BEFORE AND SIX HOURS AFTER SUNSET IN FINE WEATHER OVER LAND
The diagrams are based on data published by Lettau and Davidson.[5]

then increased with height up to 900 mb.* These vertical gradients of humidity mixing ratio allow us to use water vapour as a good tracer of vertical air motions. If, for example, we smooth the upper 1·00 g/kg line on Figure 4 to allow for errors in measurement and small-scale humidity fluctuations, we see that air which initially was at about 935 mb rose some 20 mb during the first half of the period and then sank again by the same amount. This implies vertical air motions of the order of 5 cm/s. Support for this inference comes from Figure 5. If we use entropy as the tracer we see from the shape of the 11°C isentrope that the motion implied is very similar.

* This particular distribution of the humidity is a difficult one to measure with a manual dew-point type of instrument, because as the aircraft ascends into drier air the observer may not cool the thimble quickly enough and then as the moister layer is entered the thimble will be too cold and too low a dew-point will be recorded. For this reason the apparent changes in humidity in the horizontal at the top of the inversion are not regarded as the definite indicators of an error in air position which they would otherwise have been. (On the other hand, although individual values in the regions of strong vertical gradient may be in error, there is no doubt that the strong vertical gradients were present and are correctly located.)

The coincidence of the rapid warming just at the inversion top (Figure 3) with the time of sunset is probably fortuitous, but it does seem reasonable to conclude that the cooling above 700 m before sunset was the result of adiabatic ascent and that after sunset there was a period of subsidence. If we speculate that because of the similarity with Lettau and Davidson's data, the time of the cooling may not be fortuitous and that temporary adiabatic ascent before sunset is widespread, then we must postulate a wave moving westwards with the sun. We may further speculate that the wave may be driven by the local diurnal heating over land as part of the semi-diurnal solar tide (Chapman,[6] Scorer[7]).

On this occasion the vertical motions were confined to air above the inversion. There was no vertical motion on the same time-scale at the inversion and no change of pressure at sea level comparable with the 20-mb change inferred for the air which was initially at 935 mb. We may deduce that there was a convergence/divergence sequence in air between the inversion and the 11°C isentrope, and that this was compensating (or being compensated by) a sequence in the opposite sense at higher levels.

Conclusions.

(i) Air in a 300 m deep sub-inversion layer warmed by about 0·2 degC/h as it moved over the North Sea at 70 km/h in February. Unsystematic humidity variations, equivalent to changes in the mean dew-point of the layer by ±1 degC, swamped any systematic moistening.

(ii) Gravity waves with a period of about 20 minutes were found at the inversion.

(iii) Above the inversion the air cooled before sunset and began to warm again after sunset.

(iv) Above the inversion there was adiabatic ascent before sunset and subsidence afterwards, possibly associated with a wave moving west with the sunset.

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AN EXAMPLE OF A VECTOR ERROR DISTRIBUTION DERIVED FROM A ROUTINE UPPER WIND FORECASTING TEST

By D. N. AXFORD

Introduction. The *Handbook of statistical methods in meteorology*[1] has several chapters dealing with scalar error distributions, and also deals with distributions of vector quantities, such as winds measured over a period at a particular station. It does not emphasize, however, that the frequency

distribution of vector errors is very different from that for scalars, and that quantities such as the mean and the standard deviation of the vector error distribution need to be interpreted with care, since the distribution is no longer Gaussian. Whilst statistical handbooks can be found which mention these distributions, few examples can be found.

An example of such a distribution, which arose from routine upper air forecasting tests, is presented in this note. It is shown that in this case the distribution is close to, but not exactly, a circular frequency distribution, and that the errors due to forecasting wind speed and direction are of approximately equal importance.

The forecasting test procedure. For general-purpose flying, upper air wind forecasts can be presented in the form of a map of the British Isles and surrounding area, divided into convenient boxes with a single forecast wind at each level assigned to each box.* This wind is used by navigators on any route across the box. The routine, described more fully by Singleton,[2] provides for forecasts to be made four times a day. The issue at 1000 GMT, based on the midnight contours, provides two forecasts, one valid for the period 1200–1800 GMT, and the other valid from 1800–0000 GMT. In this test the 300 mb and 200 mb winds are compared with the winds actually reported on three radiosonde ascents, Lerwick, Hemsby and Valentia. The 1800 GMT reports were used to test the earlier period forecasts (referred to as 18-hour forecasts), and the 0000 GMT reports tested the later period (24-hour forecasts).

It is clear that this system of testing is hard on the forecaster; a mid-period measurement of the mean area wind would perhaps be more valid. However, it has the advantage of simplicity and objectivity, and it had to fit into a full forecasting routine in which there was little time to give carefully weighted mean winds to large areas. From the point of view of the user these tests give the likely error for a random position within any box. They cannot be compared with mean route-wind forecasts, in which averaging will smooth the large errors to give a lower value of vector error, but they are comparable with the COMET computed area winds which are now prepared on a routine basis. Thus, tests of these results against similar computer results should give a valid comparison between forecaster and computer.

Forecasts were made over the period December 1963 to November 1964.

The vector error distribution. A vector error distribution may be considered in two ways. The normal approach used in wind forecasting practice (for example Singleton[2]), is to look at the distribution of the *magnitudes* of the vector error. Alternatively, the vector errors, quantities with both direction and magnitude, may be considered as a *vector* distribution with characteristics similar to that of the wind at a point.

Vector error distributions have been described with particular reference to wind measurements by Priestley.[3] Priestley shows that if the magnitude of the forecasting error is split into two parts, a speed error, h, and a magnitude

* Since the boxes are chosen arbitrarily the forecast winds refer to random points within the boxes.

error due to the error in forecast direction, k, and if h and k are independently subject to the normal-error law with standard errors σ_h and σ_k, then the probability of the vector error lying within the rectangle $(h_1,\ h_1+\delta h)$ $(k_1,\ k_1+\delta k)$ is

$$\mathrm{f}(h_1, k_1)\ \delta h\ \delta k \propto \exp\left(-\frac{h_1^2}{2\sigma_h^2}-\frac{k_1^2}{2\sigma_k^2}\right)\delta h\ \delta k, \qquad \ldots (1)$$

which is constant round an ellipse centred at the origin. Further, if the errors due to speed and direction are equal, so that $\sigma_h = \sigma_k = \sigma$ (say), then the frequency distribution becomes circular, such that the frequency of a total vector error of magnitude f (r) between r and $r+\delta r$ is given by

$$\mathrm{f}(r) = \frac{1}{\sigma^2}\, r \exp\left(-\frac{r^2}{2\sigma^2}\right). \qquad \ldots (2)$$

Note that this distribution has zero frequency at the origin,
a maximum or mode at $r = \sigma$,
and that the mean value is at $r = \sigma\sqrt{(\pi/2)} \simeq 1{\cdot}25\sigma$
while the root mean square (r.m.s.) is at $r = \sqrt{2}\ \sigma \simeq 1{\cdot}41\sigma$ $\Big\}$. $\ldots (3)$

The distribution f(r) described above shows the probability of obtaining vector errors of different magnitudes within a specified range. The alternative approach is to consider the distribution of the vector errors themselves. To do this the vector error $\mathbf{V}_E$ may be written

$\mathbf{V}_E = \mathbf{V}_F - \mathbf{V}_A$ where $\mathbf{V}_A$ is the actual wind and $\mathbf{V}_F$ the forecast wind.

Then the mean vector error is given by

$$\overline{\mathbf{V}_E} = \overline{\mathbf{V}_F} - \overline{\mathbf{V}_A}\ . \qquad \ldots (4)$$

Notice that the magnitude of the mean vector error $|\overline{\mathbf{V}_E}|$ is not the same as the average value of the magnitudes of the vector errors which is given by the expression $\overline{|\mathbf{V}_E|}$. The mean referred to in the equations at (3) and in Table I is the average value of the magnitudes of the vector errors.

The variance ε^2 of the vector error distribution may now be described by

$$\varepsilon^2 = \overline{(\mathbf{V}_E - \overline{\mathbf{V}_E})^2} = \overline{[(\mathbf{V}_F - \overline{\mathbf{V}_F}) - (\mathbf{V}_A - \overline{\mathbf{V}_A})]^2}$$

$$\text{or}\quad \varepsilon^2 = \sigma_F^2 + \sigma_A^2 - 2\sigma_F\sigma_A\, R, \qquad \ldots (5)$$

where σ_A, σ_F are the standard vector deviations of $\mathbf{V}_A$ and $\mathbf{V}_F$ respectively, and R is the stretch vector correlation coefficient between $\mathbf{V}_A$ and $\mathbf{V}_F$.

The vector error may be resolved into two mutually perpendicular components, for instance along x and y axes pointing to the east and to the north. Then it can be shown that the variance of the components of the vector error along these axes is given by

$$\left.\begin{aligned} \varepsilon_x^2 &= \sigma_{Fx}^2 + \sigma_{Ax}^2 - 2\,\sigma_{Fx}\,\sigma_{Ax}\,\rho_x \\ \text{and}\quad \varepsilon_y^2 &= \sigma_{Fy}^2 + \sigma_{Ay}^2 - 2\,\sigma_{Fy}\,\sigma_{Ay}\,\rho_y \end{aligned}\right\}, \qquad \ldots (6)$$

where σ_{Ax}, σ_{Fx}, σ_{Ay} and σ_{Fy} are the standard deviations of the components of the actual and forecast winds in the x and y directions, and ρ_x, ρ_y are the correlation coefficients between the actual and forecast components along the respective axes.

This approach is suitable for testing the circularity of the vector error distribution by the use of Mauchly's[4] statistic, L_e, given, in our notation, by

$$L_e = 2\,\varepsilon_x\,\varepsilon_y \sqrt{[1 - (\rho_{ab})^2]} / (\varepsilon_x^2 + \varepsilon_y^2) \qquad \ldots (7)$$

where ρ_{ab} is the correlation between the components a, b of the vector error along the northerly and easterly axes.

For an exactly circular distribution $\varepsilon_x = \varepsilon_y$ and ρ_{ab} is zero so that L_e is unity. The probability of obtaining a value of L_e less than one from a sample of N independent observations drawn from a population in which $L_e = 1$ is shown by Mauchly to be L_e^{N-2} (see Brooks and Carruthers[1] page 233). At the 5 per cent level of significance this means that the distribution can be accepted as circular provided $2/L_e < 2 \times 20^{1/(N-2)}$

$$\text{or } \log_{10}(2/L_e) \quad < (0{\cdot}7 + 0{\cdot}301\,N)/(N-2).$$

Discussion of results. The results of the wind forecasting test described previously were analysed for comparison with theory. The first part of the analysis split the magnitude of the vector error into an error in the forecast wind speed, and a magnitude error due to the incorrect forecast direction, using the formula

$$\begin{aligned} |\mathbf{V}_E|^2 &= [|\mathbf{V}_A| - |\mathbf{V}_F|]^2 + 4|\mathbf{V}_A|\,|\mathbf{V}_F| \sin^2 \varphi/2 \\ &= |\mathbf{V}_E \text{speed}|^2 + |\mathbf{V}_E \text{direction}|^2 \end{aligned} \qquad \ldots (8)$$

where $\mathbf{V}_A$, $\mathbf{V}_E$ and $\mathbf{V}_F$ are defined as before and φ is the angular error. It was hoped that resolution in this way might help to bring some of the causes of the forecasting errors to light.

Preliminary examination showed that the differences between the stations were small, with errors at Lerwick and Valentia being slightly larger than those at Hemsby, as would be expected with the sparse nature of the data over the North Atlantic. Since there were insufficient data to justify separation into significant seasonal variations, the data were combined on an annual basis into four classes defined by the two heights and the two periods of forecast.

The shape of the frequency distribution of each class is demonstrated in Figure 1 which is a composite distribution formed by totalling the errors of all four classes into one overall class. The histogram, consisting of 5-knot interval classes, is shown, and an example of a circular distribution satisfying equation (2) is also drawn based on the mean value of σ found from the distributions of the two components of the vector error. Table I summarizes the results of calculations of the mean, root mean square, and 'variance about the mean' of the magnitude of the vector error for each class.

TABLE I—STATISTICS OF THE MAGNITUDE OF THE VECTOR ERRORS $|\mathbf{V}_E|$

Height	Period of forecast	No. of observations	Mean	Root mean square	Standard deviation
mb	*hours*			*knots*	
200	24	683	22	27	15
200	18	703	19	23	13
300	24	725	31	36	19
300	18	734	26	31	17

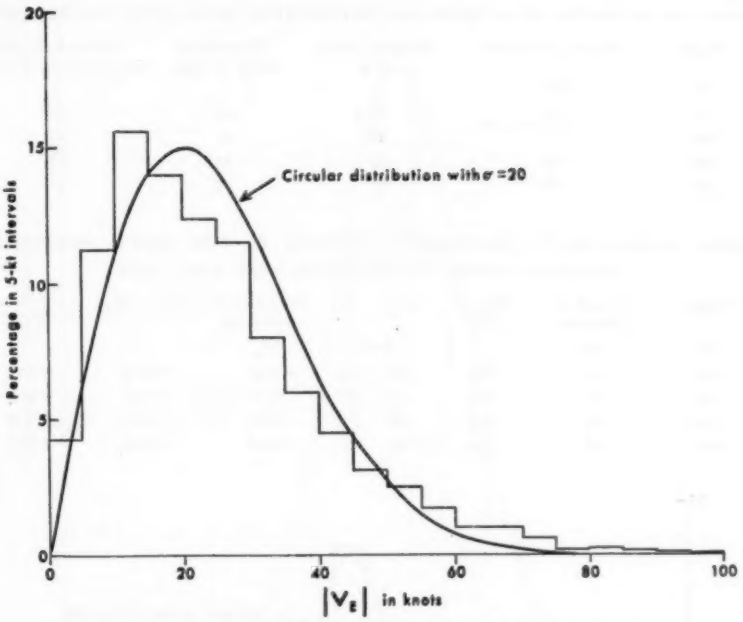

FIGURE 1—VECTOR ERROR DISTRIBUTION — ALL CASES

Figure 2 shows the distribution of $|V_E$ speed$|$ for the overall class in the form of a histogram with 10-kt interval classes. A normal curve with the same peak value is also shown for comparison.

Similarly Figure 3 shows the distribution of the errors due to the direction term for the overall class. Before constructing this graph, the distribution of the angular error φ was investigated, and found to be closely symmetric. Again a normal curve is drawn in for comparison. It can be seen that neither of these distributions is exactly normal, and, in fact, testing by means of Pearson parameters (see Brooks and Carruthers[1] page 89) confirms that the difference is significant.

It is of interest to consider the values of the mean and the root mean square which correspond to a hypothetical circular distribution in which the variances due to the speed and directional errors are equal, given the average value for each class from Table II. These can be found using equation (3) and the results are shown in Table III. These values correspond closely to those measured and shown in Table I.

TABLE II—STATISTICS OF THE ERROR COMPONENTS $|V_{E\,\text{speed}}|$ AND $|V_{E\,\text{direction}}|$

Height	Period of forecast	No. of observations	Standard deviation of $\lvert V_{E\ \text{speed}}\rvert$	Standard deviation of $\lvert V_{E\ \text{direction}}\rvert$
mb	*hours*		*knots*	
200	24	683	17	20
200	18	703	15	17
300	24	725	23	28
300	18	734	20	24

TABLE III—STATISTICS OF A CIRCULAR DISTRIBUTION WITH GIVEN VALUES FOR σ

Height	Period of forecast	Assumed value of σ	Theoretical mean ($1{\cdot}25\sigma$)	Theoretical root mean square ($1{\cdot}41\sigma$)
mb	*hours*			
200	24	18·5	23	26
200	18	16	20	23
300	24	25	31	35
300	18	22	27	31

TABLE IV—MAUCHLY'S CIRCULARITY STATISTIC L_e FOR ERROR COMPONENTS RESOLVED ALONG NORTH–SOUTH, EAST–WEST AXES

Height	Period of forecast	No. of obs.	σ_a	σ_b	Correlation coefficient	$\text{Log}_{10}(2L_e)$	$\frac{0{\cdot}7+0{\cdot}301\,N}{N-2}$
mb	*hours*		*knots*		ρ_{ab}		
200	24	683	21	19	−0·02	0·305	0·303
200	18	703	17	16	0·10	0·305	0·303
300	24	725	26	25	0·00	0·305	0·303
300	18	734	24	21	−0·09	0·305	0·303

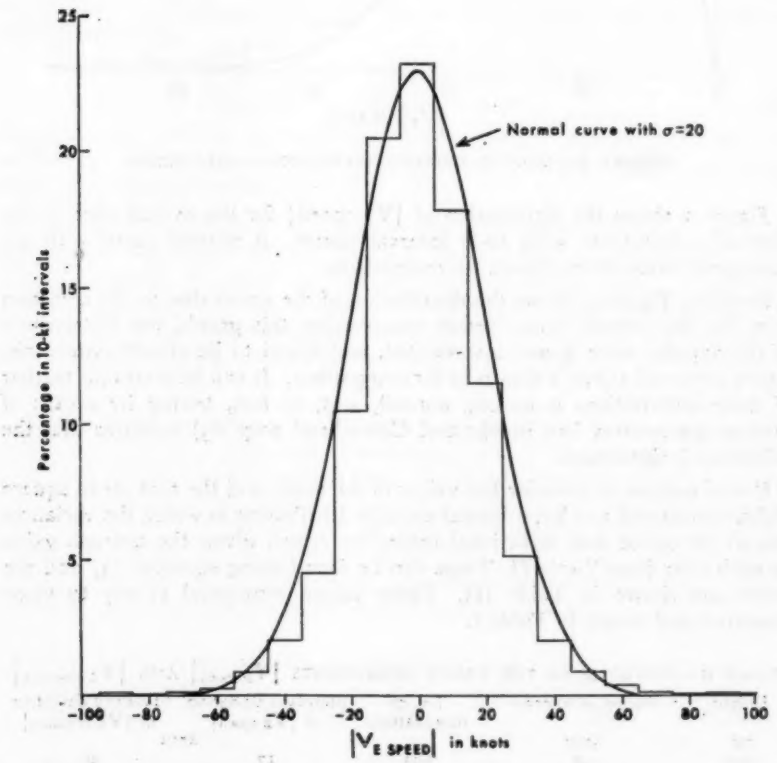

FIGURE 2—DISTRIBUTION OF ERROR COMPONENT DUE TO ERROR IN FORECAST WIND SPEED — ALL CASES

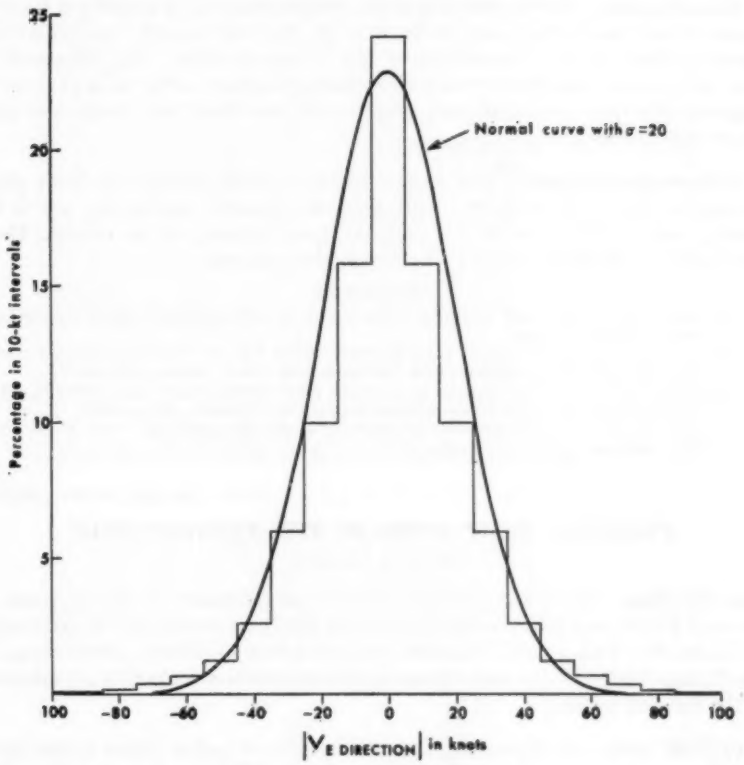

FIGURE 3—DISTRIBUTION OF ERROR COMPONENT DUE TO ERROR IN FORECAST WIND DIRECTION — ALL CASES

The second part of the analysis was designed to test for the circularity of the vector error distribution using Mauchly's L_e statistic. The vector errors were resolved along axes pointing to the north and east, and the standard vector deviations and cross-correlation coefficients were calculated for the four classes. Table IV shows the relevant values for the circularity test. The result is surprising in that in every case, at the 5 per cent level, the test for circularity fails. It can be seen that in all the classes, and in fact at every station, the standard vector deviation of the vector errors resolved in a northerly direction is greater than that in an easterly direction. A possible explanation for this difference, suggested to the author by Mr N. E. Davis, is that it is caused by the difficulty of positioning an upper trough or ridge exactly. A forecast wind of 230° 50 kt ahead of a trough when the actual wind was 310° 50 kt in the rear of the trough, gives a nil error in an easterly direction but a 65-kt error in a southerly direction. Upper troughs are nearly always orientated in a north–south direction and very seldom in an east–west direction.

Conclusions. The distribution of the vector errors in 12 months of forecast upper winds has been shown to be close to, but not exactly, a circular frequency distribution. Comparison of the histogram from 2845 observations and the circular distribution based on the appropriate value of σ (Figure 1) suggests that there are more very small errors and more very large ones than might be expected.

Acknowledgements. The author wishes to acknowledge the help given to him by Mr A. F. Crossley in bringing the Mauchly circularity test to his notice, and by Mr J. A. Harker with the programming of the results. He is also indebted to Mr S. E. Virgo for his encouragement.

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551.543.5(26.78):551.553.5

PRESSURE JUMP LINES IN THE PERSIAN GULF

By D. J. CLARK

Introduction. There have been a number of references in recent years to pressure jumps and wind oscillations in the Mediterranean, and in particular at Malta.[1-5] This article describes two occasions of similar phenomena in the Persian Gulf and discusses the main features of these and other occurrences in the last five years.

17 June 1967. At Muharraq, Bahrain, a series of pulses began in the wind speed at 0250 GMT, and lasted about an hour. The speed varied from less than 2 kt (the threshold for recording) to 18 kt in the first pulse, then in subsequent pulses to 22, 18, 18, 16, 16, 11, 11 kt before dying away completely. The wind direction exhibited no such regular oscillations as were seen on the speed trace, though rather erratic swings did occur. The traces are shown in Figure 1.

Figure 2 shows the barogram for the same day. A pressure jump can be seen at 0300 GMT; this was the subject of a 'special report' to Air Traffic Control. There was an instantaneous rise of 1·6 mb followed by a period of about an hour and a quarter when the trace was thickened to the space of the 1 mb lines.

The surface chart for 0000 GMT showed a trough from the Gulf to the Caspian Sea with a low at 50°N 50°E, and a high over north-west Europe. The air on the eastern flank of this high was probably being advected across Iraq towards the Gulf, marking a changeover in the trajectory of the air affecting this area. Unfortunately there were no observations available from Iraq, Syria and Saudi Arabia; from Iran only the observation from Tehran was received.

The time-section at Figure 3 shows the changes in the upper wind and humidity structure which occurred at the passage of the disturbance. Of

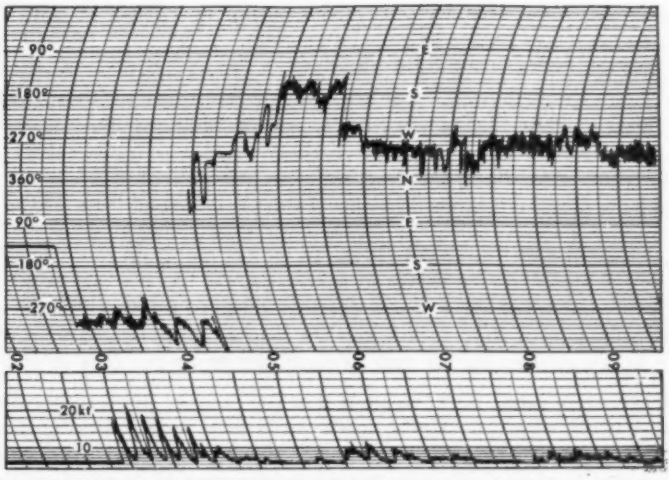

FIGURE 1—SECTION OF ANEMOGRAM FOR 17 JUNE 1967, MUHARRAQ, BAHRAIN

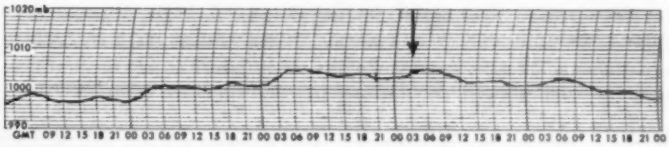

FIGURE 2—SECTION OF BAROGRAM FOR 14–18 JUNE 1967, MUHARRAQ, BAHRAIN

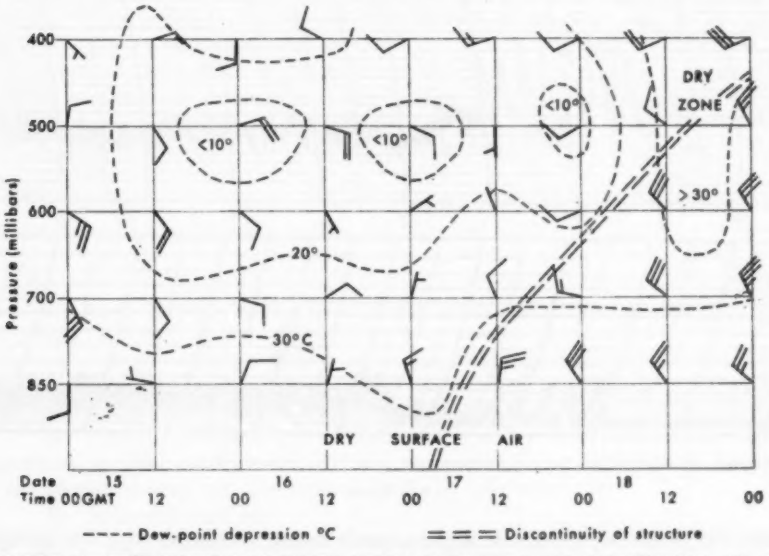

FIGURE 3—TIME-SECTION OF UPPER WINDS AND DEW-POINT DEPRESSIONS, 15–18 JUNE 1967, MUHARRAQ, BAHRAIN

particular interest is the dry zone, which may be compared with frontal structure as revealed by the series of Meteorological Research Flight measurements through fronts near the British Isles.[6]

The thermogram showed a sudden rise in temperature at about 0300 GMT of 1·5 degC and the hygrogram a rapid fall in relative humidity beginning at the same time.

At Sharjah, some three hundred miles to the east of Bahrain, at about 0500 GMT the next day, the surface wind suddenly increased from 3 kt to 15 kt, the beginning of a series of heavily damped pulses which lasted about two hours, but after half an hour (three pulses) the speed did not again reach 10 kt. The direction remained fairly constant at south-easterly. At the same time as the increase in wind speed there was a pressure jump of 0·5 mb. There was no unusual behaviour in temperature, but the hygrogram indicated a decrease in relative humidity from 84 per cent at 0400 GMT to 37 per cent at 0800 GMT.

17 September 1966. Kirk[2] suggests that the pressure jump rather than the oscillatory wind, which, at Malta, was noticed first, is the important feature; hence the name of 'pressure jump lines', and he described an occasion at Malta when a marked pressure jump was unaccompanied by wind oscillations. At Muharraq, on the other hand, there have been a number of occasions in the last four years when wind oscillations have been unaccompanied by discernible pressure jumps. One such occasion occured on 17 September 1966.

Figure 4 shows the anemogram for the early hours (GMT) of that day. The exceptionally well-marked oscillations in direction are seen to be in phase

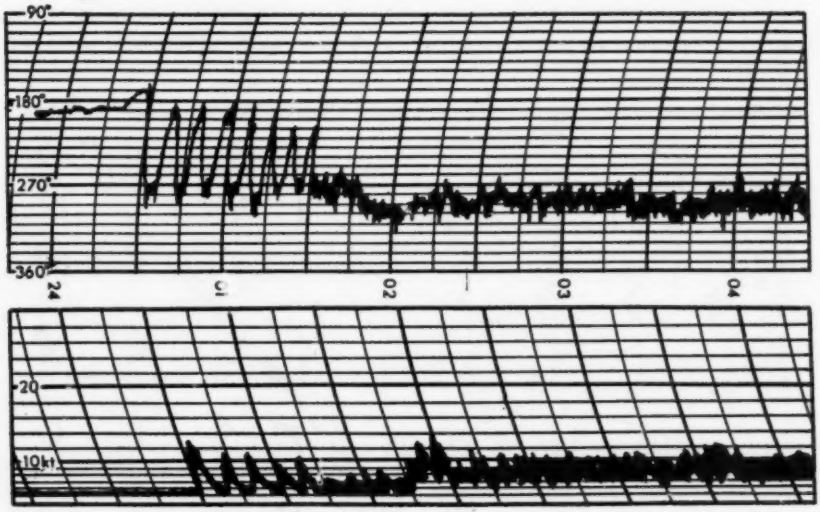

FIGURE 4—SECTION OF ANEMOGRAM FOR 17 SEPTEMBER 1966, MUHARRAQ, BAHRAIN

with the pulses in speed, the higher speeds being associated with the swings to the west. The barogram for the latter part of the week is at Figure 5. Some slight thickening of the trace may be discerned between 0000 and 0130 GMT

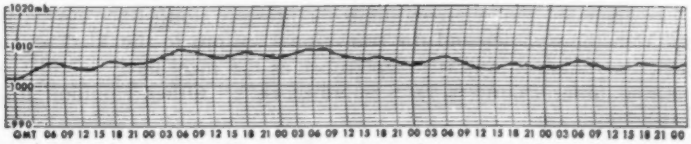

FIGURE 5—SECTION OF BAROGRAM FOR 14–18 SEPTEMBER 1966, MUHARRAQ, BAHRAIN

on the 17th, but in general the trace during the time of the oscillations presents no feature which is not paralleled elsewhere in the series of regular rises and falls. However, Sharjah experienced a very marked pressure jump some 30 hours after the Bahrain occurrence, the initial rise being of 2 mb and this was followed by a fall of 1 mb. At the same time there was a pulsing of wind speed over a period of five minutes, with variation from 11 kt to 23 kt, and an accompanying veer from 260 degrees to 280 degrees. The pulses were impressed on the sea breeze.

The 0000 GMT surface chart for 17 September 1966 showed a high pressure centre at about 50°N 70°E and a trough from Pakistan across Iran to north Iraq. There was a slack pressure gradient over the Gulf area; plotted winds were generally variable 5 kt or less, though Dhahran showed 10 kt from the north-west.

Discussion. After the September 1966 occasion of oscillatory winds the records at Bahrain, back to 1962, were examined for other examples. Twelve particularly well-marked examples were found, and there were a number of less well-marked ones. The criteria for selection were as follows :

(i) Oscillations in wind direction and speed, being in phase.
(ii) At least three complete oscillations.
(iii) A variation in wind direction of at least 40 degrees.
(iv) A final change in wind direction.

The period of oscillation was found to vary very little. Of the 12 examples, 8 had a period of about 10 minutes, 2 of 15 minutes and the other 2 of 20 minutes or more. None occurred in the months October to March inclusive : 7 occurred in the warm season from June to September, and of these 3 were in September.

The barograms (as many as were available) appropriate to each occasion were examined for pressure jumps. Of nine barograms five showed a pressure kick, though one was barely perceptible. Another showed some thickening of the trace. In the remaining three there was no discernible unusual behaviour.

Only three thermograms indicated any disturbance. One showed a fall of 0·5 degC followed by a rise of the same magnitude; another first a rise, then a fall of 1·0 degC; and the third a steep rise, then a period of about two hours with oscillations in temperature of about 1·0 degC and finally a fall to the original value.

The hygrograms were not of sufficient quality for useful examination.

It is, perhaps, significant that there are several points of similarity between the Mediterranean and the Persian Gulf. In brief these are :

(i) A land-locked sea.
(ii) The mountainous nature of much of the surrounding terrain, and the desert around most of the remainder.
(iii) An essentially two-season régime.

Kirk[3] has suggested that the conditions for pressure jump lines are especially favourable in the summer and the transitional seasons in the Mediterranean. Experience at Bahrain indicates September as the most likely month; there were several other examples in September 1967. Other common features noted at Bahrain are :

(i) A period of light winds before the oscillations begin. There is a marked preference for oscillations to appear between 2100 and 0600 GMT (8 out of the 12) when winds are usually at their lightest but this probably merely indicates that stronger winds conceal the oscillations.

(ii) The disturbance heralds a change in air-mass characteristics, though in general not associated with 'weather'. However, on one occasion a line of stratus moved across the airfield at Muharraq, becoming cumuliform as it did so. As it passed over, the wind flicked round to north-west and a series of oscillations began. Later in that day there was a cumulonimbus cloud over Qatar, east of Bahrain.

Kirk[3] associates the pressure jump line with the movement of ridges and troughs in the high atmosphere. In the Persian Gulf data are too scanty to justify the association. He also suggests that the pressure jump line might be included as a recognized feature in analysis.[2,3] In the Persian Gulf area, however, it appears to be of interest rather than significance; there have been no reports of low-level turbulence, for example, as described by Grimmer[7] and commented on by Kirk.[5] In any case, since on most plotted charts the immediate neighbourhood is represented by a sliver of observations along one side of the Gulf, the prospects of being able to incorporate these interesting phenomena in an analysis are slim indeed.

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551.509.311:551.577.37(424)

AN EXCEPTIONALLY HEAVY RAINFALL IN JULY 1968

By P. R. S. SALTER

Introduction. In the 24 hours ending 0000 GMT, 11 July 1968 more than 100 mm of rain fell in parts of Gloucestershire from thunderstorms associated with a small depression which had moved north-eastwards from Spain into the south-west of England. This depression then moved east-north-eastwards, deepening rapidly as it crossed East Anglia. Further heavy rain fell in areas adjacent to the Wash.

The object of this article is to trace the sequence of events which led to the rainfall. A mechanism for the severe thunderstorms is suggested.

The rainfall. Figure 1 shows the approximate rainfall distribution during the period when the depression was close to southern England, i.e. the 48 hours ending 1800 GMT, 11 July, (11/1800 GMT). It may be useful to compare this rainfall with previous heavy rainfalls in the west country in recent times.

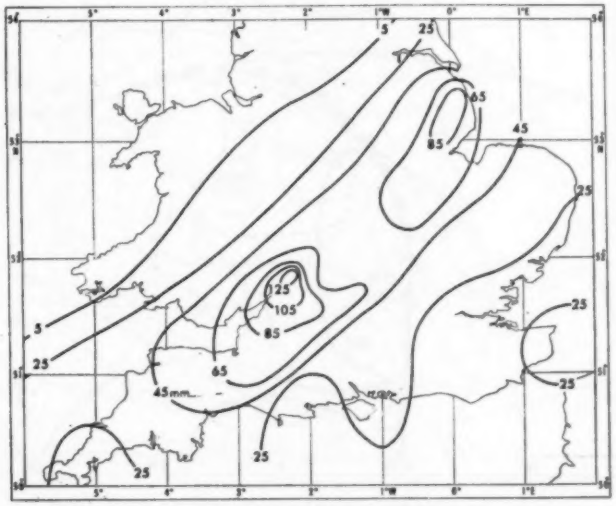

FIGURE 1—RAINFALL FOR 48 HOURS ENDING 1800 GMT, 11 JULY 1968
Isopleths at 20-mm intervals

In August 1952, more than 200 mm of rain fell in 24 hours on parts of Exmoor to cause the Lynmouth disaster. In July 1955, over 270 mm of rain fell within 24 hours in a storm at Winterbourne Steepleton, Dorset. What was so significant about the rainfall in July 1968 was the size of the area which received the heavy rain. In Figure 1 the area of the 105-mm isopleth is nearly 1000 km^2. Over an area of 1000 km^2, 100 mm of rain represents 10^{11} kg of water.

The surface synoptic situation. The surface synoptic situations for 09/1800 GMT and 11/0600 GMT are illustrated in Figures 2 and 3 respectively. The surface wet-bulb potential temperatures (hereafter called WBPT) for the earlier time have been indicated in Figure 2. The significance of these temperatures will become apparent later.

On 8 and 9 July unstable medium-level cloud was observed in parts of France but no thunderstorms were reported in France until after 09/1800 GMT. This lack of thunderstorms was important in the final analysis. The warm front depicted in Figure 2 was moving slowly northwards bringing up air which in France had a dry-bulb temperature of around 27°C with a dew-point temperature of 20°C. In the western Mediterranean air temperatures reached 35°C and locally 40°C in Spain. The pressure began to fall slowly in France during the afternoon of the 9th, an event which together with the presence of unstable medium-level cloud often precedes thundery weather.

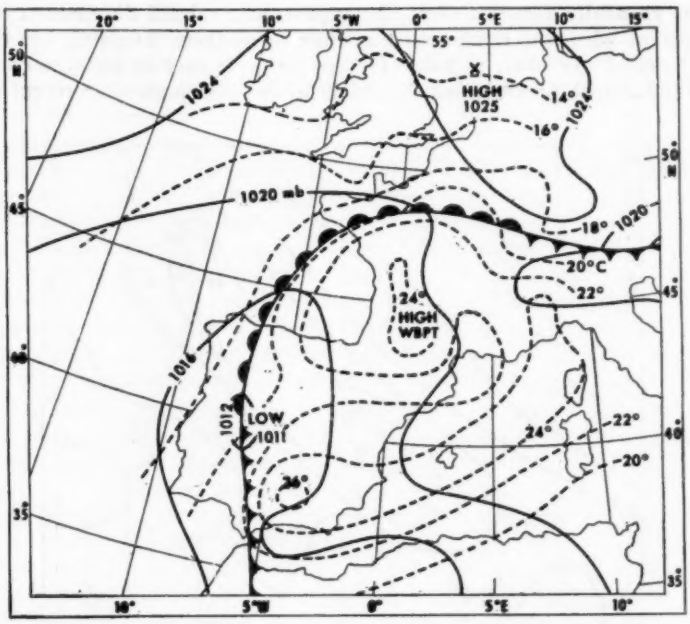

FIGURE 2—SURFACE CHART 0600 GMT, 11 JULY 1968
Isobars at 4-mb intervals
Surface wet-bulb potential temperature at 2 degC intervals

The severe thundery activity which affected the south-western counties, and also south coastal regions initially, moved into those regions during the night of the 9th/10th. For much of the 10th the thunderstorms were restricted to the south-west while the surface low of 1006 mb slowly approached from the Bay of Biscay. Soon after 10/1800 GMT the depression began to move east-north-eastwards towards East Anglia and to deepen rapidly. The associated warm front crossed southern counties earlier in the afternoon with heavy rain at the frontal passage. The thunderstorms, which were distributed near the middle of and also ahead of the surface low, spread across central southern England and East Anglia into the Low Countries. The early hours of the 11th found the depression with a central pressure of 994 mb in the southern North Sea and still deepening — Figure 3 *(a)* refers to the situation at 11/0600 GMT. Northerly winds reached 60 kt over the sea while gale force winds were recorded over parts of East Anglia. Much rain fell in counties adjacent to the Wash with some thunderstorms occurring in the northerly flow behind the depression which then moved into north-west Europe leaving the whole of England and Wales in an unstable north-westerly airstream with pressure rising steadily everywhere.

The upper air synoptic situation. The upper air pattern prior to 10 July had been progressive over the southern Atlantic near the Azores. The charts at 300 mb are indicated in Figures 3*(b)* and *(c)* for 09/1200 GMT and 11/0000 GMT respectively, together with the positions of the maximum winds

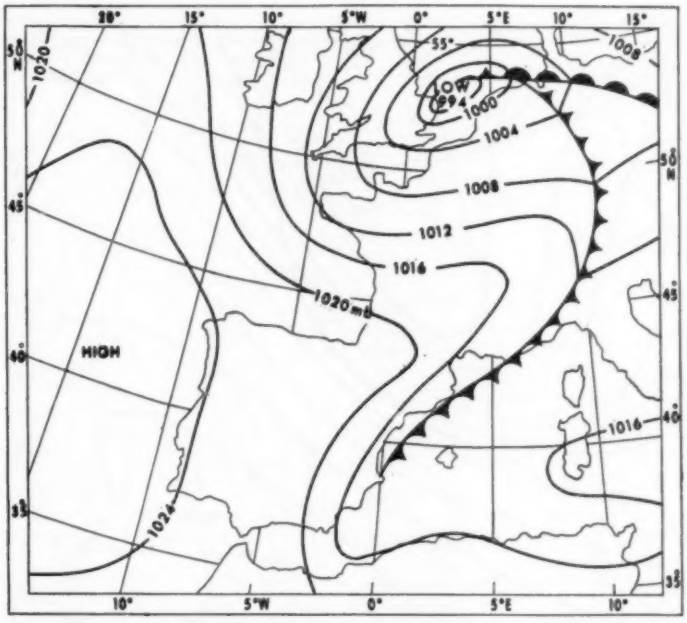

FIGURE 3*(a)*—SURFACE CHART 0600 GMT, 11 JULY 1968
Isobars at 4-mb intervals

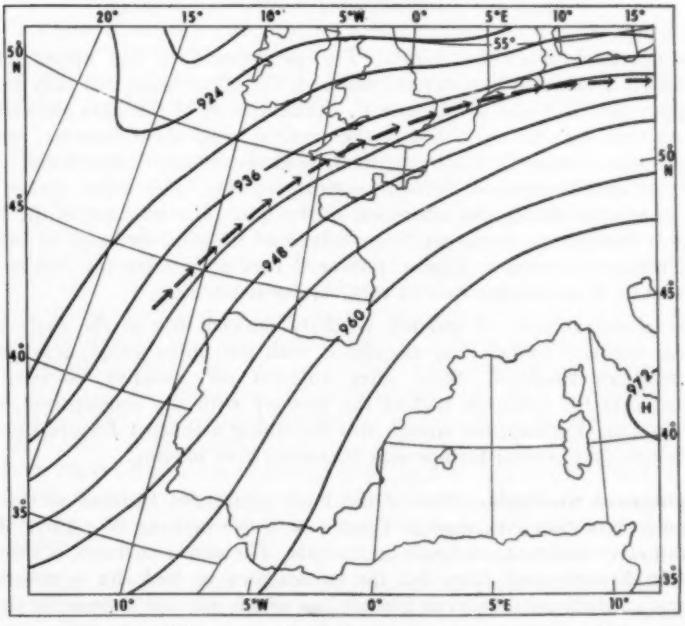

FIGURE 3*(b)*—300 mb CHART 1200 GMT, 9 JULY 1968
Isopleths at 6-decametre intervals. → → → Jet axis

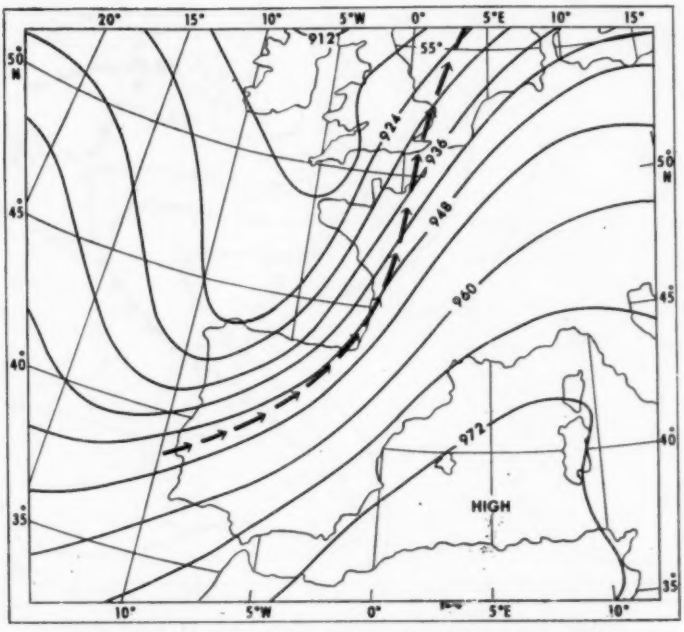

FIGURE 3*(c)*—300 mb CHART 0000 GMT, 11 JULY 1968
Isopleths at 6-decametre intervals. → → → Jet axis

which occurred around 250 mb. The progression of the upper trough, particularly in the south, was most marked, the effect being not only to back the upper flow but also to increase it. Examination of the data showed that from 09/1200 GMT to 10/0000 GMT the vertical wind shear between 500 mb and 300 mb increased. Strong vertical wind shears have been found[1] to be important mechanisms for self-propagating storms, and these shears were being generated during the afternoon of the 9th. The backing of the upper flow also resulted in warm air being advected towards the area of interest. The Camborne ascents in Figure 4 showed that at 850 mb the temperature rose from 04°C at 09/1200 GMT to 10°C twelve hours later.

A graphical analysis of 300 mb vorticity showed that in the early stages negative vorticity (which was associated with the upper ridge) was present over southern England. Soon after 10/1200 GMT positive vorticity was advected into the southern half of the country with the acceleration of the upper trough. Further, the trough was becoming a sharper feature since the upper ridge over central Europe was becoming slow moving.

Tephigram analysis. One of the most significant features of this case was the lack of thunderstorms in France until the evening of 9 July. There were thundery outbreaks in Spain on the 9th. The non-occurrence of thunderstorms in France until later led the investigator to look for a restraining influence on deep convection in France but which was not necessarily present in Spain.

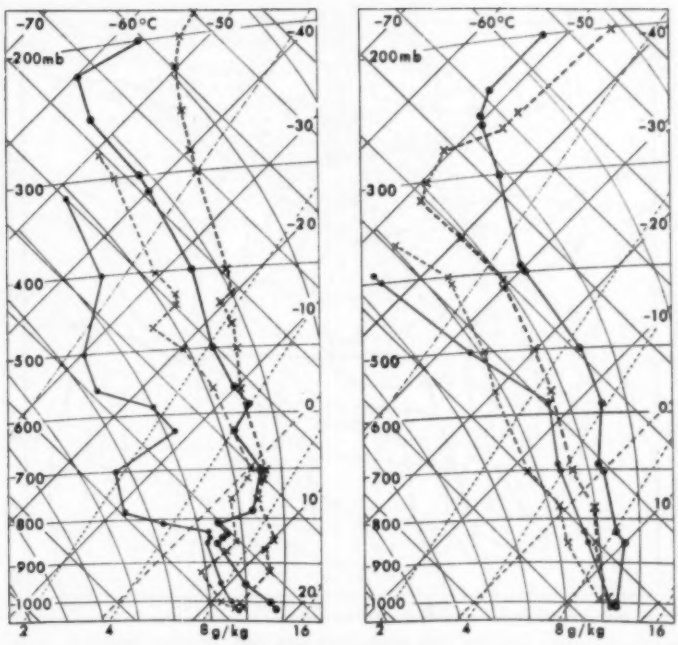

FIGURE 4—CAMBORNE ASCENTS 9–11 JULY 1968

Temperature 1200 GMT, 9 July 1968
Dew-point 1200 GMT, 9 July 1968
Temperature 0000 GMT, 10 July 1968
Dew-point 0000 GMT, 10 July 1968
Temperature 1200 GMT, 10 July 1968
Dew-point 1200 GMT, 10 July 1968
Temperature 0000 GMT, 11 July 1968
Dew-point 0000 GMT, 11 July 1968

Examination of the surface WBPT at 09/1800 GMT (see Figure 2) showed a maximum of 25°C in the lee of the Pyrenees there being a large area of 23°C in France. These values of surface WBPT were abnormally high (compared with an average value in France in summer of 14°C), and were found by an examination of a time sequence of charts to have been advected from the Mediterranean. Figures 5*(a)* and *(b)* show the advection which occurred between 08/1200 GMT and 10/0000 GMT. Analysis of the dry-bulb potential temperatures showed that at 850 mb. and to a lesser extent at 700 mb, an area of potentially very warm air was present over parts of Spain and France. Figures 5*(a)* and *(b)* show this 'Spanish plume'[2] of warm air in relation to the surface WBPT. When the surface air temperature reached 35°C, as in parts of Spain, then deep convection proceeded. In France, however, maximum temperatures reached only 30°C on the 9th which was not sufficiently high to release instability by surface heating alone. Thus convection in France was minimal because there was a plume of potentially warm air which effectively acted like a lid.

The manner in which deep convection was released can be seen from Figure 5*(b)* to be the escape of the high surface WBPT from beneath the plume of warm air. This escape was caused by backing of the surface gradient which

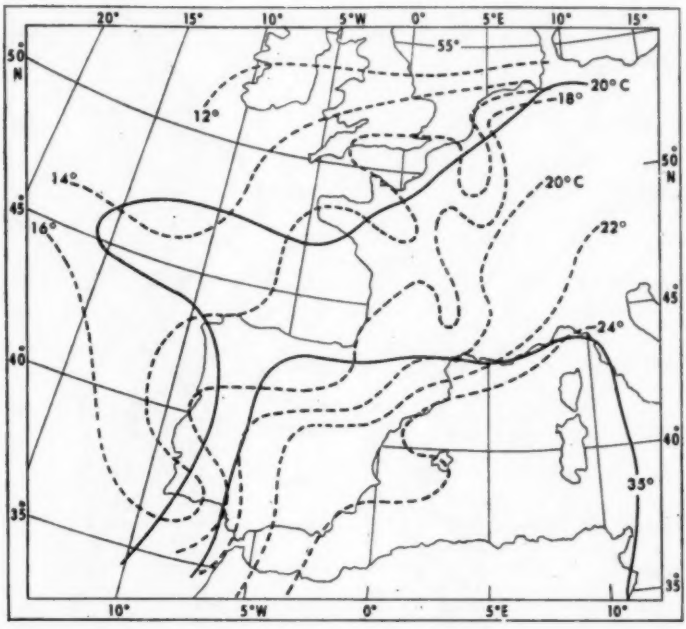

FIGURE 5*(a)*—MOIST AIR AT THE SURFACE AND DRY AIR AT 850 mb, 1200 GMT, 8 JULY 1968

– – – – Surface wet-bulb potential temperature ——— 850 mb potential temperature

resulted from a fall in the surface pressure field; the fall was initially due to diurnal variation. Once the warm moist air in the lowest levels of the atmosphere had emerged from the restraining influence of the plume then convection could proceed to beyond 300 mb. The process was further enhanced by the increasing strength of the upper flow.

The Camborne ascents, in Figure 4, showed certain interesting features. Between 09/1200 GMT and 11/0000 GMT there was an increase in the moisture content between 800 mb and 400 mb. At the same time warming occurred in the upper troposphere as a result of the backing upper flow advecting warmer air from the south. By 10/1200 GMT the Camborne ascent showed that colder air from the west was beginning to penetrate the sounding and at 11/0000 GMT the cold air had completely penetrated with the arrival of the upper trough.

Instability indices. Since 10 July was an occasion of severe thunderstorms in the south-west of England it seemed pertinent to test techniques of instability forecasting. Saunders[3] has tested thunderstorm forecasting techniques for eastern England and found that of the methods available those due to Rackliff[4] and Boyden[5] were often the most useful. Both these indices were plotted for western Europe throughout the synoptic period of the 9th–11th, and found to over-estimate the possible area of thundery activity when used completely as objective tests. A third index, the Showalter index,[6]

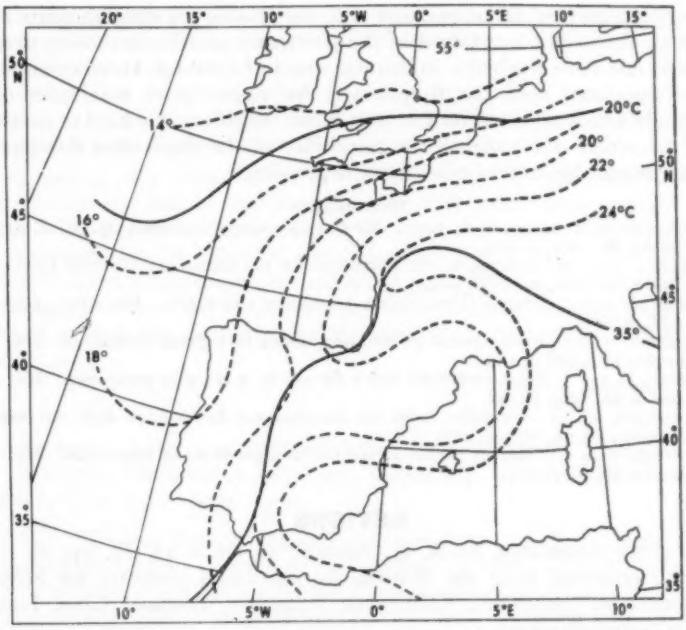

FIGURE 5*(b)*—MOIST AIR AT THE SURFACE AND DRY AIR AT 850 mb, 0000 GMT, 10 JULY 1968

– – – – Surface wet-bulb potential temperature ——— 850 mb potential temperature

which has been used in the U.S.A. for forecasting thunderstorms and tornadoes, was tested in the same way as the other indices and was found to be more realistic. However any instability criteria should be used in conjunction with all the meteorological factors attendant to a synoptic case.

Benwell[7] has written of the possibility of using the jet stream at 500 mb as a predictor of heavy rain. When used on the data for 10 July, the criterion was marginally satisfied in southern England, so that heavy rain might have been expected there provided a correct forecast was available for the upper winds required in Benwell's method.

Discussion. Thundery situations in summer time have often in the past proved difficult to forecast, both regarding the distribution of thunderstorms and their severity. Certain techniques for forecasting thunderstorms have been tried in this synoptic study with limited success. They did not give an accurate indication as to the severity of the thundery activity which produced such heavy rain, particularly in south-western England, on 10 July. On that day a complex meteorological situation occurred with firstly, severe thunderstorm activity and secondly, rapid deepening of the depression as it crossed eastern England. Examination of the data has suggested that the method of release of instability over England was the advection of air of high surface WBPT from beneath the 'Spanish plume', and that the vertical wind shears were the means of maintaining the storms for a considerable time.

The deepening of the depression was due mainly to developments in the upper air flow. The movement of the jet stream and its increasing strength, which at one time reached a maximum speed of over 140 kt, were important factors associated with the deepening. An upper wind maximum of the magnitude which occurred on the 10th is comparable to the kind of maximum found in winter and suggests the possibility of the depression developing a vigour comparable to that of a winter depression.

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REVIEWS

Theory of fog condensation, by A. G. Amelin. 25 cm × 18 cm, pp. xi + 236, *illus.*, (translated from the Russian by the Israel program for Scientific Translations, Jerusalem), Oldbourne Press, 1–5 Portpool Lane, London, E.C.1, 1967. Price 99*s*.

The theory of dropwise condensation is presented in this book from first principles. The mathematical derivations are given in considerable detail and so are very easy to follow. The theoretical findings are supported by full relevant experimental data. The book is not intended primarily for the meteorologist and examples of the effects of fog formation during industrial processes abound throughout.

First there is an introductory chapter in which such basic concepts as critical supersaturation, homogeneous condensation, condensation on nuclei and accommodation coefficients are described. Having explained the importance of supersaturation, the next five chapters are devoted to the mechanisms of the formation of supersaturated vapour and fog by distinctive means. Chapter two deals with adiabatic expansion and radiative cooling, and is mainly a thorough description of the principles of the expansion cloud chamber. The section on fog formation by radiative cooling is very short and superficial.

The third chapter is concerned with fog formation by turbulent mixing of gases and provides a good description of fog formation in turbulent jets (not the meteorological kind). Next the diffusion cloud chamber is described as an example of supersaturation produced by molecular diffusion and thermal conduction. Chapter five, the longest in the book, is concerned with mists formed in turbulent flow. The formation of fog during vapour condensation on a surface is described in detail. The last part of the chapter describes means of preventing fog in such common industrial processes as the recovery of solvents by condensation, freezing out of vapour, and condensation in cooling towers.

Of the remaining two chapters, the first describes the formation of mists as the result of chemical interaction between gaseous minor constituents. Included is a short section on the formation of atmospheric nuclei by this mechanism. The book is concluded with a short description of methods of removing unwanted mists in industry and also of the production of controlled mists for use in agriculture, medicine and research.

After reading this book one is left with the strong impression that the author is certainly not a meteorologist. He uses some surprising examples of natural fog and cloud formation to illustrate the various modes whereby supersaturation can be achieved. For instance frontal cloud is used as an example of cloud formation by turbulent mixing. However, the fundamental facts of fog formation are on the whole well presented, and the book serves to remind us that fogs and mists are not the sole prerogative of the meteorologist.

P. GOLDSMITH

The Environmental Science Services Administration, by Roy Popkin. 22 cm × 14 cm pp. x + 278, Frederick A. Praeger, Publishers, 77–79 Charlotte Street, London, W.1, 1968. Price: 50*s*.

The U.S. Weather Bureau, Coast and Geodetic Survey and Central Radio Propagation Laboratory were grouped together in 1965 in the U.S. Department of Commerce to form a single unit, the Environmental Science Services Administration (ESSA). The author is concerned with showing why this grouping was considered necessary. He starts with an historical account of each from its governmental inception to the merger itself, shows the reasons for believing that the combined manpower and instrumental facilities could be deployed more profitably, describes the structure of ESSA and finally considers its role in the immediate future. Much of the information is drawn from official sources so that the book provides an essay into the way in which governments make decisions about their scientific policies.

The main interest for meteorologists will be in the structure of ESSA and the plans for the near future, especially concerning international co-operation and warning services for severe weather conditions. There is little information about the research institutes which are so well known to us, perhaps because the services that are available have more interest for the general reader.

E. KNIGHTING

LETTERS TO THE EDITOR

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Some uses of profiles and charts of total energy content

In his article,[1] Dr J. G. Lockwood makes the important claim on page 147 that 'in particular, Q profiles indicate the presence of large-scale ascent'.

The ability to recognize from a radiosonde ascent whether or not ascent is occurring would be of such enormous diagnostic and prognostic value that it behoves us to examine how this can be done from a Q profile when it cannot be done from a T-φ gram.

Dr. Lockwood implies in two statements in the second paragraph of page 150 that there was ascent in the layer of nearly constant Q at two soundings. Now what does constancy of Q with height mean in terms of a T-φ gram ? It occurs :

(a) With a dry adiabatic lapse rate and a dry atmosphere.

(b) With a wet adiabatic lapse rate and saturation.

(c) With a lapse rate less than wet adiabatic and dew-point depression increasing at such a rate with height that the lapse rate of wet-bulb potential temperature is zero.

(a) is not a real case except in the upper troposphere. The fact that *(b)* is the distribution that would arise if an air mass with the characteristics in *(c)*, i.e. zero lapse of wet-bulb potential temperature, were lifted, does not really allow us to infer ascent from it. The air may have been lifted earlier, and in any case air masses undergoing vigorous convection over a warm sea often show this distribution.

Moreover, soundings taken in rain areas ahead of warm fronts frequently show a lapse rate considerably less than the wet adiabatic over a deep layer: ascent is occurring in a place where there is a positive lapse of wet-bulb potential temperature. Such soundings would show a marked increase of Q with height.

There are many ways in which a study of the distribution of Q can be rewarding, but I think that attempts to deduce vertical motion from the Q profiles are likely to be misleading.

Meteorological Office, Bracknell

M. K. MILES

Reply by Dr J. G. Lockwood :

In my article I suggested that by reasoning from first principles, it is likely that a constant vertical Q profile indicates ascent at the time of observation or in the recent past. Mr Miles suggests several ways in which constant vertical Q profiles could arise. The first example does not occur in the troposphere. The second example could arise by ascent or possibly in several other ways, none of which are likely to be of importance in the free atmosphere except perhaps the evaporation of falling rain. In general water vapour is supplied to the atmosphere from below and transported upwards by convection, the second distribution is therefore likely to imply ascent now or in the recent past. As soon as ascent ceases the Q profile will be modified by radiation losses and wind shear. The third example is unlikely to occur over deep layers in the free atmosphere.

Intense rainfall often occurs in non-frontal troughs or in intense convergence lows. Systems of this type should show constant vertical Q profiles indicating intense ascent. In an active warm front, warm air overlies cold air, and rain falls from the warm air through the cold air thereby causing high humidities in the cold air. Soundings through systems of this type will show a positive lapse of wet-bulb potential temperature through the frontal inversion. In an active warm front, ascent is taking place in the warm air above the frontal inversion, and it is here that a layer with constant Q should be found. Q will increase across the actual frontal surface.

Ascent is possible with a marked increase of Q with height, but it is unlikely for the following reasons. Such a system would be energy-consuming, and it would require an outside energy source, therefore it is unlikely to become self-generating. Q values increasing with height during ascent suggest that air with continually decreasing Q values is being brought into the system at the surface. This cannot continue for a long time. Also decreasing Q values at the surface imply falling temperatures and humidities, hence less available energy to drive the system.

I agree that Q profiles cannot by themselves indicate the rate of ascent, but I suggest that they can indicate if marked ascent is taking place or has taken place in the recent past.

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1. LOCKWOOD, J. G.: Some uses of profiles and charts of total energy content. *Met. Mag., London*, **97**, 1968, p. 147.

CORRIGENDUM

Meteorological Magazine, October 1968, p. 304.

Under Summer (ii), line 4, the significance figure 1·0 should be 10.

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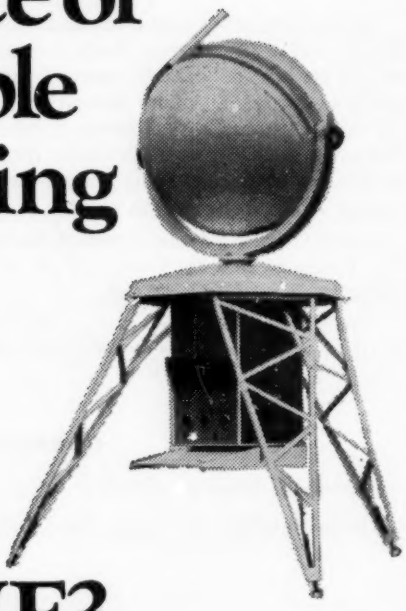

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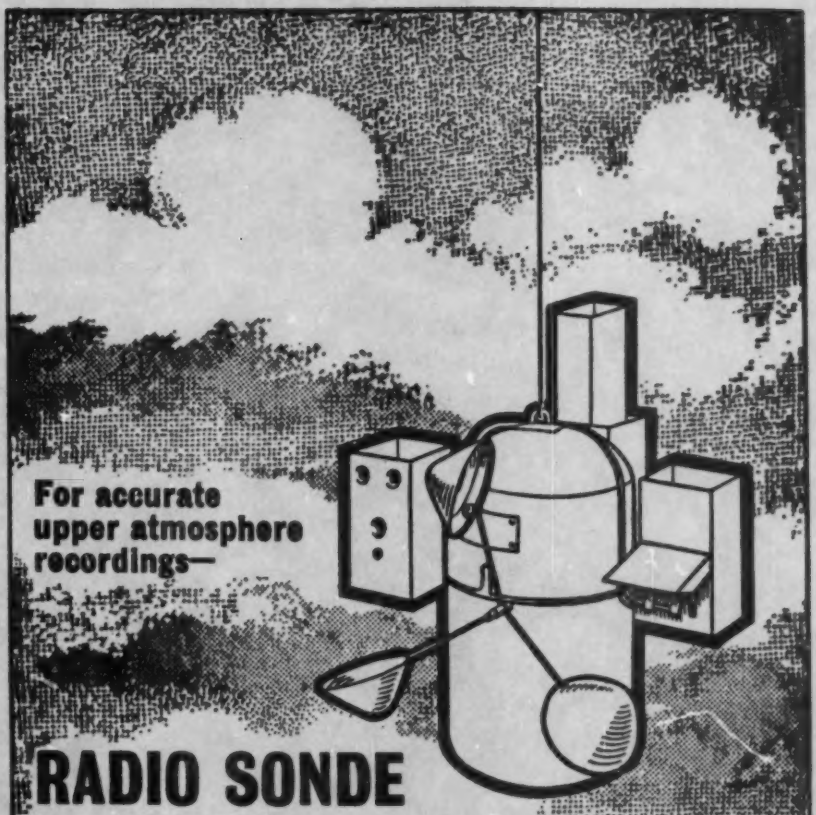

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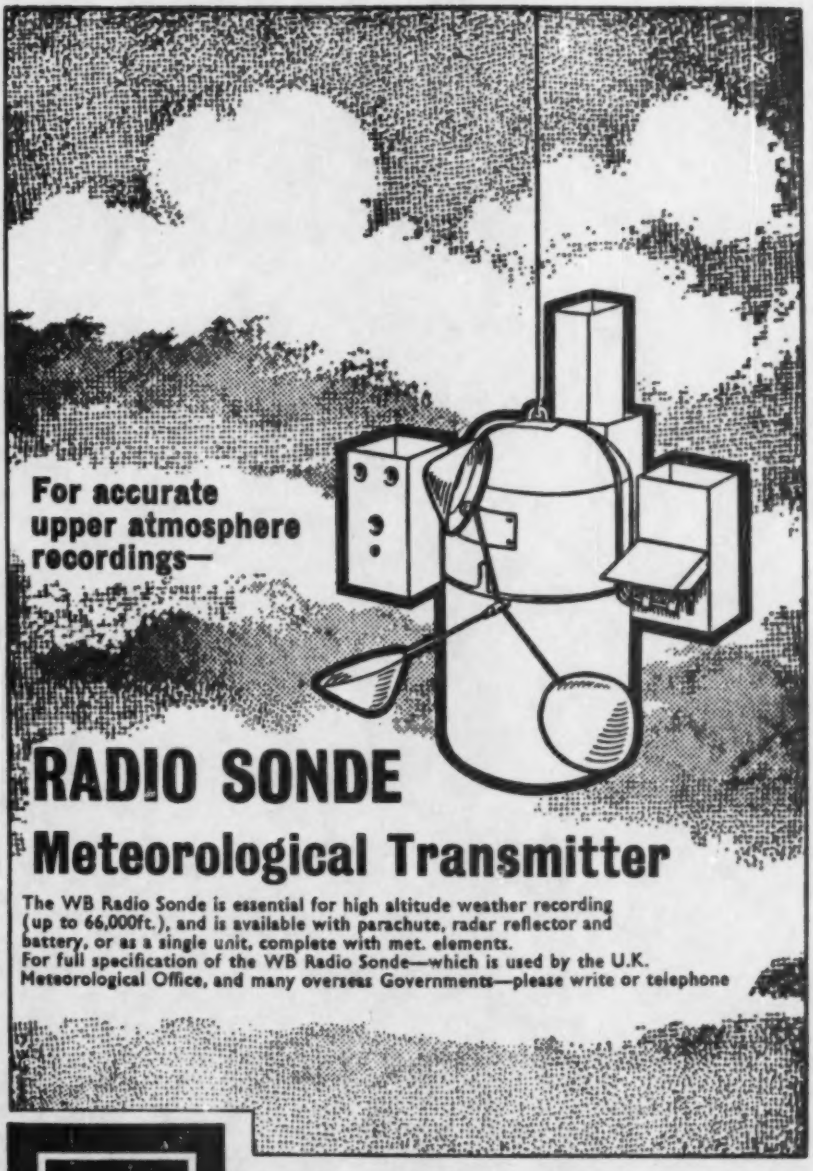

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